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Stratospheric temperatures in Antarctic winter: Does the 40-year record confirm midlatitude trends in stratospheric water vapour?

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SUMMARY

Water vapour is a potent greenhouse gas, and the observed increases in water vapour in the stratosphere act to cool it. Possible changes in stratospheric temperatures are important for future ozone loss because colder temperatures in the edge region of the Antarctic ozone hole act to increase polar stratospheric clouds there, and so delay recovery of the ozone hole. Trends in lower-stratospheric temperature within the core of the Antarctic vortex in winter should be a unique indicator of trends in stratospheric water vapour, because neither changes in CO_2 nor in ozone have a large effect on temperature in the lower stratosphere in the dark. Here, measured stratospheric temperatures southward of 70°S in winter are presented and their quality and corrections discussed. The character and magnitude of the long-term changes at Halley (76°S) are similar from 100 to 70 hPa and at 50 hPa, whether corrected for sonde changes or not, and are also similar to those at other Antarctic sites. We found no significant trend in temperatures at Halley between 1960 and 2000, which is inconsistent with the change calculated from the trend in lower-stratospheric water vapour in northern hemisphere midlatitudes between 1960 and 2000. Over the shorter interval between 1980 and 2000 at Halley, the change in temperature was -1.8 ± 0.6 K, in agreement with the change calculated from the trend in stratospheric water vapour in northern hemisphere midlatitudes between 1980 and 2000. The differences between these periods are discussed in terms of: possible fortuitous agreement between 1980 and 2000; the poorer representation and quality of the measurements of stratospheric water vapour between 1960 and 1980; and a possible large variation in the rate of oxidation of CH_4 to H_2O in the upper stratosphere between 1960 and 1980. Such a variation in oxidation rate was observed by satellite between 1992 and 1999.

KEYWORDS: Brewer–Dobson circulation Greenhouse effect Lower stratosphere Ozone

1. INTRODUCTION

The mixing ratio of water vapour in the lower (150 to 50 hPa) and middle (50 to 15 hPa) stratosphere at midlatitudes has been increasing significantly for several decades. Between 16 km and 28 km above the central USA, Oltmans *et al.* (2000) measured trends of about 1% year^{-1} between 1980 and 2000 (values varied from 0.7 to 1.4% year^{-1}, 1-sigma errors varied from 0.1 to 0.3% year^{-1}). There was a similar upward trend between 18 and 26 km above the eastern USA between 1964 and 1976, although the scatter in the measurements was larger. In the lower stratosphere above the UK, a similar trend occurred between 1954 and 1976 (SPARC 2000; Rosenlof *et al.* 2001). Other instruments (HALogen Occultation Experiment (HALOE), Atmospheric Trace Molecule Spectroscopy, MkIV radiometer) measured similar trends in the middle stratosphere more globally though over shorter time periods (SPARC 2000).

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Tropospheric air enters the lower stratosphere in the tropics, where the cold temperatures near the tropopause dehydrate it, and then it moves upwards and polewards. In the upper stratosphere (15 to 1 hPa), water vapour is produced by oxidation of methane (CH_4, hence the sum $2CH_4 + H_2O$ is approximately conserved throughout the stratosphere); but little air in the lower stratosphere at midlatitudes contains extra water vapour from CH_4 oxidation.

Hence the trend measurements cited above suggest that the entry value of water vapour into the tropical lower stratosphere has been increasing steadily at about 1% year^{-1} since 1954. Although the full controls on this entry value are in dispute (Dessler 1998; Holton and Gettelman 2001), it is clear from the agreement in phase of the annual cycles in tropopause temperatures and lower stratospheric water vapour (Mote *et al.* 1996) that the temperature of the tropical tropopause helps in this control, with colder temperatures giving smaller water vapour amounts. However, both the meteorological tropopause and the cold point have been getting colder since the 1960s (Chakrabarty *et al.* 2000; Seidel *et al.* 2001; Zhou *et al.* 2001), so that there is an apparent conflict with the increasing trend in water vapour.

In the upper stratosphere, Smith *et al.* (2000) showed from HALOE measurements that the trend in water vapour between 1992 and 1997 was more than 2% year^{-1}, but became about 1% year^{-1} when measurements to 1999 were included. The trend in upper stratospheric CH_4 was slightly negative between 1992 and 1997, contrasting with the continuing increase in tropospheric CH_4. The resultant trend in $2CH_4 + H_2O$ between 1992 and 1997 in the upper stratosphere approximately equalled 1% year^{-1} plus the trend in $2CH_4$ in the troposphere (Pyle *et al.* 1999). These observations would be consistent with a slowing of the Dobson–Brewer circulation between 1992 and 1997 which then reversed, as a slower circulation should give more time for CH_4 to oxidise to H_2O (Pyle *et al.* 1999).

Apart from effects on the chemistry of the stratosphere, any increase in water vapour will affect stratospheric and surface climate because water vapour is a potent greenhouse gas. The radiative forcing at the surface due to a 20% increase in stratospheric water vapour is a modest 0.2 W m^{-2} (Shindell 2001). However, the effects on stratospheric temperature are more pronounced—any increase in stratospheric water vapour acts to cool the lower and middle stratosphere because the increased water vapour emits more infrared radiation to space. A decrease in ozone in sunlight acts to cool the stratosphere, because less ultraviolet and visible sunlight is absorbed, slightly offset by the small decrease in infrared radiation to space (small, because the infrared bands of ozone are mostly opaque from the lower stratosphere). Trends in CO_2 barely affect temperatures in the lower stratosphere because the infrared bands of CO_2 are almost completely opaque there.

Hence the observed midlatitude trends in both water vapour (increasing) and ozone (decreasing) act to cool the lower stratosphere. Using a climate model which includes a well-defined stratosphere, Forster and Shine (1999) calculated a global midlatitude lower stratospheric cooling since 1980 of about 1 K due to the change in water vapour, and a further 1 K due to the change in ozone, in total rather larger than the observed value of a little over 1 K (Pyle *et al.* 1999).

In Antarctica in winter, the change in temperature calculated by Forster and Shine (1999) due to the change in ozone is small ($<$0.3 K) because of the lack of sunlight. In the vortex core in winter (the core being the comparatively well-mixed region at potential-vorticity (PV) equivalent latitudes southward of about 70°S) there is little mixing of air from sunlit regions in the edge region of the vortex (Lee *et al.* 2001) or from outside the vortex. In early winter the vortex is often comparatively symmetric,

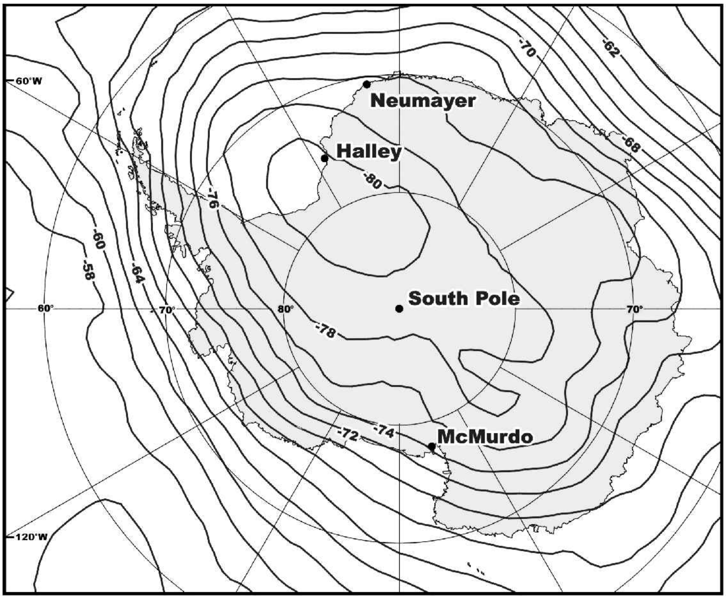

Figure 1. Map showing European Centre for Medium-Range Weather Forecasts analysed Antarctic temperatures (°C) in early winter (8 June 2000) at 100 hPa, illustrating the size and typical distribution of the vortex, together with the locations of radiosonde sites used in this study.

as illustrated by the −72 °C contour in Fig. 1, so that in early winter there is little exposure of air in the vortex core to sunlight (in spring the vortex core is often much more elliptical and more offset from the Pole). Hence, changes in temperature of the lower stratosphere in the vortex core in midwinter should be a unique indicator of changes in water vapour, dynamical factors being constant. This contrasts with changes in temperature in the edge region of the vortex in winter or in the vortex core in spring, where in each case there is enough sunlight and/or enough ozone loss to confuse the diagnosis.

In this paper, we discuss the measurements of lower-stratospheric temperatures south of 70°S in Antarctic winter: their availability, their quality, their possible correction for changes in sonde type, and their medium and long-term changes. To discriminate against changes due to changes in CO_2 (Forster and Shine 1999, 2002) we focus on temperatures at 100 hPa. To discriminate against changes due to changes in ozone we exclude temperatures in August from our winter months, because sunlight returns to the lower stratosphere southward of 70°S from late July. We discuss the size and significance of the observed changes, fit trend lines, and discuss their implications in terms of our understanding of the size and effect of trends in stratospheric water vapour.

We also focus on changes at 100 hPa because this is also a useful altitude for discriminating against changes in dynamical heating. Whilst dynamical heating due to descent in the winter vortex must occur at all altitudes, the largest changes should occur at altitudes above 10 hPa in Antarctica. Shindell *et al.* (1997) found dynamical heating of over 2 K at 4 hPa in October, induced by an ozone hole in their model, but less than 0.5 K at 100 hPa. Similarly, Mahlman *et al.* (1994) found that an ozone hole in their model induced a warming of 6 K at 3 hPa, but less than 1 K below 10 hPa. By doubling CO_2 in their model, Rind *et al.* (1998) found a change in dynamical heating in winter of over 1 K above 40 km, but between zero and −0.5 K below 22 km.

TABLE 1. YEARS OF RADIOSONDE LAUNCHES AND SONDE TYPES AT THE ANTARCTIC SITES USED IN THIS WORK

Site name and WMO station number	Location	Inclusive dates of sonde data used	Type of sonde	Inclusive dates when this sonde type flown
Halley, 89022	75.6°S, 26.8°W	1957–2000	UKMO Whiteley MkII	1957–71
			Graw M60	1971–75
			VIZ1207 and 1395	1975–84
			Vaisala RS80	1984–92
			AIR 4A and 5A	1992–99
			Vaisala RS80	2000–02
South Pole[1], 89009	90.0°S	1961–2000	not known	1960–61
			VIZ	1961–94
			AIR	1994–99
			AIR 1680	1999–2000
McMurdo, 89664	77.9°S, 166.7°E	1957–2000	not known	1960–92
			VIZ	1992–95
			AIR	1996–97
			Vaisala RS80	1997–99
Neumayer, 89002	70.7°S, 8.3°W	1983–2000	Graw M60	1985–92
			Vaisala RS80	1993–2000

[1]South Pole station is formally designated 'Amundsen-Scott'.

It is vital to study trends in water vapour in the Antarctic lower stratosphere and possible causes of the trends, because of their potential to delay recovery of the ozone hole (Roscoe and Lee 2001). Recovery is expected during the next 50 years, as halocarbons are reduced due to the provisions of the Montreal Protocol for the Protection of the Ozone Layer. The potential to delay recovery arises because the edge region of the Antarctic ozone hole is isolated from its core (Lee *et al.* 2001), so that colder temperatures in the edge region would act to increase polar stratospheric clouds there. This would increase the proportion of chlorine compounds in the edge region converted to the reactive forms that take part in ozone loss.

2. MEASUREMENTS

As discussed in section 1, to isolate changes in stratospheric temperature which are due to changes in stratospheric water vapour, as opposed to those due to changes in other radiatively active gases, we analyse temperatures in the lower stratosphere near 100 hPa in June and July, south of 70°S. The four available sites with long and continuous series of radiosonde balloon flights in winter are listed in Table 1, together with details of sondes used. Of other possible sites, Dome Fuji (77.3°S) had only one year of winter data, and data in our archive for Vostok (78.5°S) only covered 1980 to 1991 (some Vostok data since July 1957 exists via a commercial website, but four continuous years of winter values are absent and there is no description of sonde types).

Many studies of trends examine monthly means, and adopt World Meteorological Organization (WMO) rules to insist on ten or more values, with no gap of more than 4 days, for a valid monthly mean (Met Office 1979). This avoids the potential for bias when there is a trend during the month, and the possibility of increased scatter with a small sample. We have adopted these rules for means of single months (which we use below only for illustrative purposes), except for the data-quality subtractions in Table 2 and the corresponding Fig. 2 where it is obviously inappropriate to delete any values (see below).

TABLE 2. HALLEY (75.6°S, 26.8°W) 100 hPa TEMPERATURES: MEANS AND STANDARD DEVIATIONS OF THE DIFFERENCES (K) BETWEEN THE MONTHLY MEANS OF FLIGHTS WITH VALUES ABOVE 100 hPa AND OF THE MONTHLY MEAN OF ALL FLIGHTS

Month	Mean	Standard deviation
June	+0.10	0.43
July	+0.12	0.55
August	+0.08	0.36

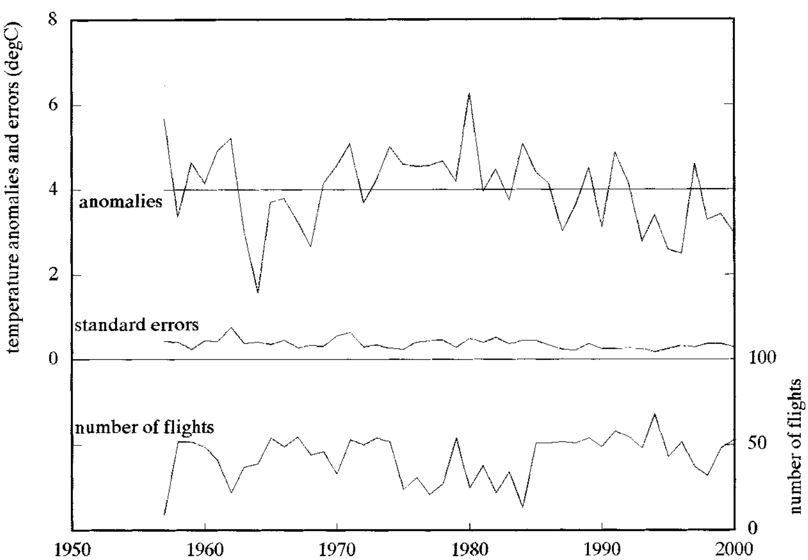

Figure 2. Anomalies of average temperatures in June + July at 100 hPa measured by radiosondes at Halley (75.6°S, 26.8°W) from their 1957 to 2000 mean, together with the standard error of each 2-month average (left-hand scale), and the number of flights in each 2-month period (right-hand scale). Anomalies are shifted upward by 4 degC for clarity. Note that errors are consistent with the year-to-year scatter in the anomalies, and that the error does not in general increase when there are fewer flights.

Other results in the text, and in the figures and tables below, are the means of the two adjacent months June and July combined (subsequently June + July). The trend over this 2-month period is large, about 10 K, so that even with an adaptation of WMO rules it would be essential to remove the bias due to incomplete sampling. We remove the bias as follows: first the mean day number in each 2-month sample is found, where day numbers run from 1 on 1 June to 61 on 31 July; then we find the mean temperatures in June and in July, each averaged over all the years in the dataset, and their difference divided by 30.5 is taken to be the mean daily trend in temperature during June and July for the whole dataset; the bias to be subtracted is then the mean day number in each sample (minus 30.5) multiplied by the mean daily trend.

In the Halley June + July means, the minimum number of days is nine, in 1957 which is not used in trend fits. The minimum number used in trend fits is 13 days (see Fig. 2). We tested to see if the variability was larger in smaller samples by calculating the standard error in each 2-month period (after correcting for the temperature

Figure 3. Average temperatures in June + July at 100 hPa measured by radiosondes at Halley (75.6°S, 26.8°W) from 1957 to 2000, with no smoothing (upper trace), and with various smoothing filters. Values are corrected for changes in sonde types. Unfiltered values for 2001 are included in the filtering. Filtered traces are offset downwards by 2, 3.5, 5 and 6.5 degC for clarity and ease of scaling. For later plots, we chose a triangular filter of half-width 3 years (centre trace) as a useful compromise between retaining the medium-term features of unfiltered data and reducing the scatter.

trend during the period, otherwise the trend would artificially inflate the error estimates). The peak-to-peak variability in Fig. 2 is about four times the r.m.s. standard error, illustrating the consistency of our approach. Figure 2 confirms that larger standard errors have little correlation with smaller numbers of days, except for one year (1962). We feel that this further justifies our not using an adaptation of WMO rules for our means of two adjacent months. Finally, in straight-line fits to the Halley data below, we test fits weighted by the number of flights in the 2-month period, and there is little difference from unweighted fits.

In data from McMurdo and South Pole, we rejected two-month means with fewer than 3 days. We justify this small sample because the means are only used to provide qualitative support for our findings, and because a significantly larger rejection threshold would have excluded much of the McMurdo data and significant parts of the South Pole data.

Although the resultant scatter in the mean June + July values from year to year at Halley is modest, it is large enough to justify some smoothing to help guide the eye and make it easier to discern any medium and long-term changes. We experimented with several smoothing functions for presentational purposes, illustrated in Fig. 3. Our choice for all the other plots in this work was a triangular function of full width at half maximum (FWHM) 3 years, achieved by two successive 3-year running means. Gaps of three or more years were left as gaps in the plots (there were no gaps in Halley data at 100 hPa), but gaps of one or two years were treated by allowing the first running mean to average any years in its window (this gives a function which is a truncated or distorted triangle but still about 3 years FWHM). At the ends of each series, the end and next-to-end values are excluded from the plots of smoothed data.

3. Data quality

From Halley, temperatures were archived at the British Antarctic Survey (BAS), both at significant levels (turning points in the temperatures during the ascent) and at standard levels (100 hPa, 50 hPa, etc.). Quality was ensured by the normal checks for range and continuity.

From other sites, since 1987, many temperatures were archived in near real time at BAS, after being transmitted over the WMO's Global Telecommunications System. Some were received later as part of the REference Antarctic Data for Environmental Research (READER) archiving project of the Scientific Committee for Antarctic Research (SCAR). Only values at the standard levels were archived, applying quality control similar to that for Halley data.

An important question regarding data quality for trend analysis is: do balloons burst more readily when they are colder? The answer is widely known to be 'yes' for some types of untreated balloons, because of their increase in brittleness when the balloon is colder, due to a combination of low air temperatures and, if dark, the lack of solar heating of the balloon. If balloons do burst more readily when colder, it would bias the results by reducing any apparent trend.

At Halley, we guard against the possibility of increased brittleness by dipping balloons in an 8:1 mixture of aircraft fuel and oil for 30 minutes, and then draining for 30 minutes before inflation. Since 1995, balloons are also kept at 60 °C for 48 hours before dipping. These procedures had been found empirically to be effective with the Totex types TA or CR balloons that have been used at Halley for many years.

Nevertheless, there is still a case to answer, particularly as data from Halley since the switch to 350 g balloons in 1992 is frequently absent above about 30 hPa. We investigated the Halley data for signs of more bursts at lower altitude in colder conditions, by comparing winter temperatures at 100 hPa when values were present above this level and when they were not. If absent, then the balloon had burst just above 100 hPa. If balloons did burst more often just above 100 hPa when it was colder at 100 hPa, and temperatures are correlated with those just above 100 hPa, then the mean temperatures at 100 hPa when values were present above would be warmer than when values were absent above.

The results in Fig. 4 and Table 2 show that the differences between these mean temperatures were negligibly small (less than 0.15 K), were not statistically significant (less than a quarter of the standard deviation), and were smallest in the coldest month (August). Hence the frequent absence of data above 30 hPa since 1992 can be ascribed solely to the switch to smaller 350 g balloons, the more so as a burst-altitude of 30 to 20 hPa is consistent with what we would expect from the smaller balloons.

There have also been many changes in sonde types. Archived values had no radiation corrections until the automated receivers introduced by Vaisala in 1980. At night, corrections depend only on the infrared emissivity of the sensor screen, and on its ventilation which is proportional to the pressure. Differences between many sondes were carefully measured in an intensive intercomparison reported by Nash and Schmidlin (1987), who also listed the time constants of the sensors. Radiation corrections were also calculated by Luers and Eskridge (1998). Table 3 lists the various corrections which can be made with current data, although there are some gaps in the archived data on sonde types. The major gap is in our knowledge of sonde types at McMurdo, hence we have been unable to make corrections to McMurdo data for sonde types.

Figure 5 shows the result of applying corrections for sonde types to the Halley data; there is little difference to the medium- and long-term changes in temperature.

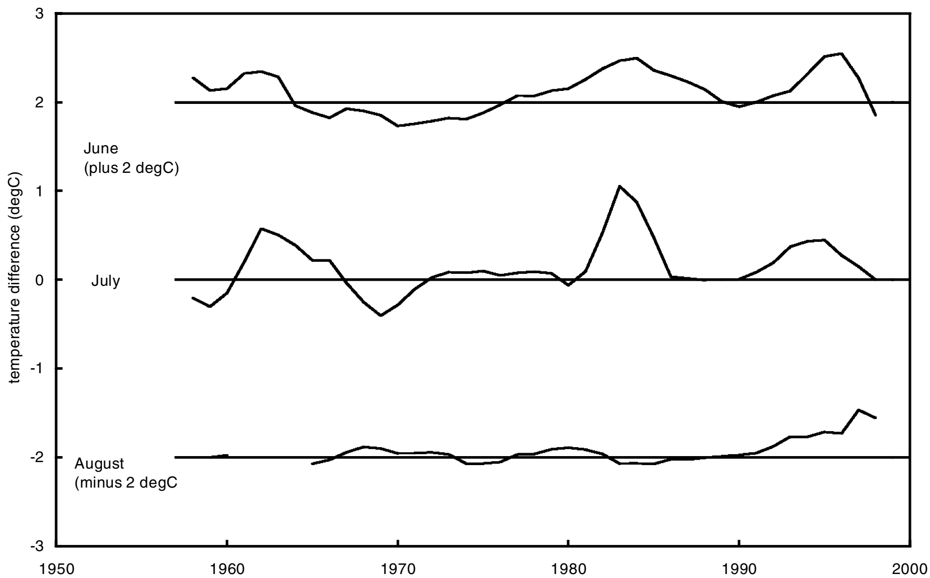

Figure 4. Differences between monthly average temperatures in winter at 100 hPa at Halley (75.6°S, 26.8°W) using only flights with data above 100 hPa, and using all flights. June and August data are displaced upward and downward, respectively, by 2 degC for clarity. Values are smoothed by a triangular filter of half-width 3 years. Missing data indicate fewer than three flights in that month. Note that the temperature-scale spans 6 degC, whereas in other figures except Fig. 9 it spans 12 degC.

TABLE 3. CORRECTIONS WHICH SHOULD BE MADE TO ARCHIVED 100 hPa NIGHT-TIME TEMPERATURES MEASURED BY DIFFERENT SONDE TYPES

Sonde type	Radiation correction (K)	Time-lag (s)	Time-lag correction[1] (K)	Total correction (K)	Reference for radiative and time-lag corrections
Vaisala (1980 to 86)	−0.89	7	−0.11	−1.00	Nash and Schmidlin (1987) Luers and Eskridge (1998)[2]
Vaisala (1987 to 93)	−0.14	7	−0.11	−0.25	Nash and Schmidlin (1987) Luers and Eskridge (1998)[2]
Vaisala (1993 to 2001)	0.00	7	−0.11	−0.11	Nash and Schmidlin (1987) Luers and Eskridge (1998)
Graw M60	−0.25	18	−0.29	−0.54	Nash and Schmidlin (1987)
VIZ and AIR[3]	+0.25	7	−0.11	+0.14	Nash and Schmidlin (1987)
UKMO Whiteley MkIIb	0.00	14	−0.23	−0.00[4]	Scrase (1956)

[1]We calculate that the average lapse rate for the whole Halley dataset between 100 hPa and 70 hPa was 0.15 K hPa^{-1}, equal to about 2.7 K km^{-1} (colder aloft); at a typical ascent rate of 6 m s^{-1}, a time-lag therefore adds 0.016 K s^{-1} to the true temperature.

[2]Vaisala sensors have very small infrared emissivity but Vaisala had incorrectly inserted a radiation correction into their automated software in 1980, equal to 0.75 K at 100 hPa (Nash and Schmidlin 1987). This was replaced by a smaller value in 1987 but not by zero until 1993 (Luers and Eskridge 1998).

[3]AIR used VIZ rod thermistors, with the same radiation correction at night and the same time-lag, although the AIR mounting arrangement differed so that the correction in daylight differed (J. Nash, private communication).

[4]Procedures at BAS sites using this sonde were to manually displace the measured temperatures to the pressures measured earlier, following UK Met Office rules, so the hand-written temperatures included time-lag correction.

Figure 5. Average temperatures in June + July at 100 hPa at Halley (75.6°S, 26.8°W), smoothed by a triangular filter of half-width 3 years, with and without corrections for sonde types (see Table 3). The corrected trace (lower) is displaced downwards by 4 degC for clarity. The main features in the medium and long term are similar.

4. RESULTS, TRENDS AND DISCUSSION

Discerning trends or patterns in data where the variability is similar to the size of any trend is fraught with difficulty. There are many advanced statistical techniques that might help convince the sceptic, but it is more convincing if similar trends or patterns are seen in independent subsets of the data or with several analytic treatments.

With this in mind, we show the Halley data at three different altitudes in June + July and for the two months separately at 100 hPa, in Figs. 6 and 7; and in Fig. 8 we show the data from the four available sites. Temperatures from McMurdo appear to have larger variability, associated with frequent sampling of air in the edge region and outside the vortex (as evidenced by their being the warmest in the dataset, and consistent with the offset nature of the vortex displayed in Fig. 1). Nevertheless, most of these datasets have striking similarities:

(i) little if any trend over the whole dataset;
(ii) a maximum in the late 1970s;
(iii) a steady decrease of about 2 K since the late 1970s;
(iv) a dip of about 2 K in the 1960s;
(v) a dip of about 1 K in 1994.

Figure 9 shows the extreme change in temperature in November since the start of the ozone hole, now over 10 K, and shows that the features in winter data listed above have a quite different character, as well as amplitude, to those from ozone loss later in the spring.

To determine trends we chose the two periods used by Forster and Shine (2002), who calculated the change in temperature due to a globally uniform trend in water vapour, which they set equal to the observed trend in water vapour in the middle stratosphere at northern hemisphere midlatitudes. These periods were 1960 to 2000 and

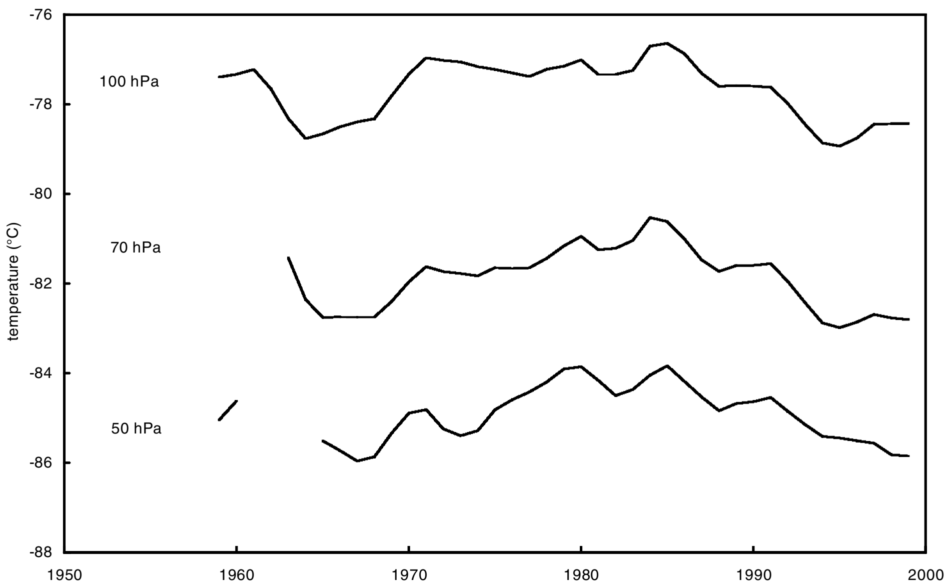

Figure 6. Average temperatures in June + July at Halley (75.6°S, 26.8°W) at 100, 70 and 50 hPa, smoothed by a triangular filter of half-width 3 years. Values at 70 hPa were not recorded until after 1961, the gap in the 50 hPa trace shows absent data. Note the similarity of medium and longer-term features at each pressure.

Figure 7. Monthly averaged temperatures at Halley (75.6°S, 26.8°W) at 100 hPa in June and in July, smoothed by a triangular filter of half-width 3 years. Note the similarity of some features: decrease since the late 1970s; minima in the 1960s and in 1994; and near-zero trend over the whole time series.

Figure 8. Average temperatures in June + July at 100 hPa at four sites south of 70° S (see Table 1 for details), corrected for changes in sonde types (except McMurdo) and smoothed by a triangular filter of half-width 3 years. Note the similarity of features at three of the four sites in each case: decrease since the late 1970s; minima in the 1960s and in 1994; negligible trend from early 1960s to late 1990s.

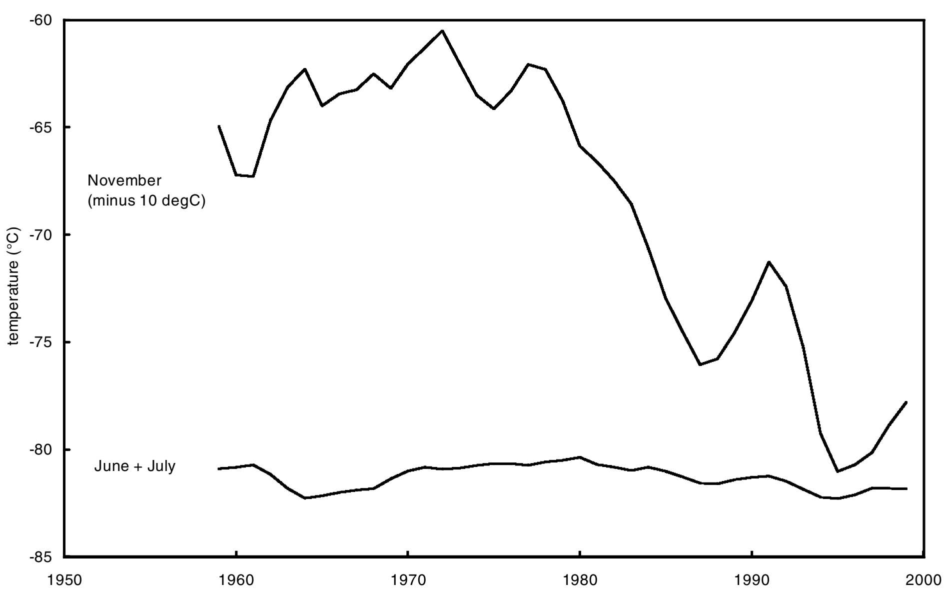

Figure 9. Monthly averaged temperatures at Halley (75.6° S, 26.8° W) in winter (June + July) and late spring (November) at 100 hPa, smoothed by a triangular filter of half-width 3 years. The November values have been displaced downwards by 10 degC to allow a finer temperature-scale. The small temperature changes in winter discussed in this paper bear little resemblance to the amplitude or style of the changes caused by the sunlit ozone hole in November, with temperatures in the 1990s 8 to 15 degC lower than in the 1960s and no minimum in the mid-1960s.

TABLE 4. TRENDS AND ONE-SIGMA ERRORS IN TRENDS (K), AT HALLEY (75.6°S, 26.8°W) AT 100 hPa AVERAGED OVER JUNE AND JULY COMPARED TO CALCULATED CHANGES FROM FORSTER AND SHINE (2002, THEIR FIG. 3)

Location	1960 to 2000	1980 to 2000
Halley, measured, unweighted	-0.6 ± 0.5	-1.8 ± 0.6
Halley, measured, weighted	-0.6 ± 0.5	-1.6 ± 0.6
75°S, calculated	-4.2	-1.6

Values are corrected using the data in Table 3. Weightings are in respect of the number of flights in the two-month period.

1980 to 2000, chosen because the observed trend is continuous and well defined since 1980 (Oltmans *et al.* 2000), whereas before 1980 it is less well defined except in the lowest stratosphere and its continuity is less clear (SPARC 2000; Rosenlof *et al.* 2001). We computed least-squares fits to the Halley measurements using standard regression formulae (e.g. Topping 1972). In order to ensure that years when there were few flights in the 2-month period did not bias the fits, we also calculated fits with weights equal to the number of flights in the period (i.e. assuming that the variance is proportional to the number of flights in each sample). The results in Table 4 show that weighted and unweighted fits give similar trends and error bars, and they confirm the results that were clear by eye in (i) and (iii) above—the observed changes between 1960 and 2000 disagree completely with the calculations by Forster and Shine (2002), despite apparent agreement between 1980 and 2000.

One possible explanation for this disagreement is that the water vapour and temperatures are unrelated, and the agreement between 1980 and 2000 is fortuitous; but, given the body of theory suggesting the link between water vapour and temperature, this seems unlikely. Another possible explanation is that the eccentricity of the vortex could have changed, such that less extra-vortex air was sampled in the 1960s and 1970s, resulting in colder temperatures then; but, given the similar behaviour at all sites, this also seems unlikely. A rigorous investigation of trends in vortex behaviour from the European Centre for Medium-Range Forecasts re-analysis data would be a major research project beyond the scope of this work.

Another possible explanation is variation in volcanic aerosol which acts to warm the lower stratosphere. However, the signature in the temperature record due to aerosol from Mt Pinatubo, which arrived above Antarctica following the vortex breakdown in November 1991, is very small or absent. We might expect an effect on winter temperatures starting in 1992 and decreasing with the observed initial time constant of 1 ± 0.2 years (WMO 1999), but the observed temperatures at Halley show no sign of a warming pulse in 1992, in either the unfiltered data (Fig. 3) or in the smoothed data (Fig. 5), unless it is obscured by the cooling pulse in 1993–94 due to changes in CH_4 oxidation (see below).

We also calculated the trend in optical depth at 550 nm, of volcanic aerosol at 30 to 90°S, tabulated by Sato *et al.* (1993) and extended to 2000, to be $-0.015 \pm 0.0.011$ optical depth units between 1960 and 2000. This is a change in optical depth over the whole period of $-16 \pm 13\%$ of the maximum optical depth of Pinatubo. Not only is this small, but it is slightly negative because of the disproportionately large amounts of aerosol in the southern hemisphere from the eruption of Mt Agung in 1963. Hence it corresponds to a slight cooling, rather than the warming needed to explain the discrepancy between measurements and calculations in Table 4.

A more probable explanation relates to the use in calculations, at all latitudes, of trends in water vapour observed at midlatitudes in the middle or lower stratosphere. Air arriving at the lower stratosphere in Antarctica in winter originates in the middle and upper stratosphere at midlatitudes. At midlatitudes, only the trend in the northern hemisphere middle stratosphere is well established, and that only since 1980.

Furthermore, as discussed in section 1, much of the water vapour in the upper stratosphere derives from CH_4 oxidation. HALOE has demonstrated that the trend in water vapour in the upper stratosphere was large and variable between 1992 and 1999 (Smith *et al.* 2000), but the trend in $2CH_4 + H_2O$ was approximately constant and about equal to the 1% year^{-1} of the lower and middle stratospheric water vapour (Pyle *et al.* 1999). As also discussed in section 1, this suggests that the rate or amount of oxidation of CH_4 to H_2O increased then slowed, possibly caused by circulation changes in response to stratospheric aerosol from Mt Pinatubo, as also suggested by the model study of Considine *et al.* (2001). It is noticeable that the dip in Antarctic stratospheric temperatures in the mid 1990s (point (v) above) occurred simultaneously with HALOE measurements of a larger trend in water vapour in the upper stratosphere.

This leads us to speculate that the mean trend in water vapour at midlatitudes in the upper stratosphere from 1980 to 2000 was similar to that observed in the middle stratosphere (1% year^{-1}), but that the mean trend in water vapour at midlatitudes in the upper stratosphere from 1960 to 1980 was much smaller, possibly due to circulation changes from an unknown cause.

One other effect that could reduce trends might be a reduction in local water vapour concentration by the formation of polar stratospheric clouds (PSCs). This could provide local negative feedback—colder temperatures would reduce water vapour locally by condensation, countering the trend of increased water vapour. However, this requires the ability to increase PSC cover, whereas PSCs within the vortex core are already ubiquitous (Cacciani *et al.* 1997), particularly in August. The 2-month thermal time constant at 100 hPa would also discriminate against such local feedback.

The discussion in the previous four paragraphs about trends in water vapour in the lower stratosphere in Antarctic winter, given trends elsewhere or other factors, would be greatly simplified if there were any actual measurements of trends in Antarctic winter, but unfortunately these are lacking. Of possible measurements: HALOE and POAM* are solar occultation sensors so cannot observe in the dark of Antarctic winter; the MLS† measured H_2O only in the middle stratosphere and only for 2 years; and Lyman-alpha sondes have only been regularly launched from McMurdo since the mid 1990s.

5. Conclusions

Changes in temperatures at 100 hPa in the lower stratosphere in the core of the Antarctic vortex in winter since the late 1950s are similar to those at 70 and 50 hPa, whether corrected for changes of sonde types or not, and regardless of whether all sites are included or only Halley.

At Halley, the trend between 1980 and 2000 after correction for sonde changes is -1.8 ± 0.6 K, but the trend between 1960 and 2000 is -0.6 ± 0.5 K. This is consistent with calculations using the observed trend in stratospheric water vapour between 1980 and 2000, but not with calculations between 1960 and 2000. One possible explanation is that large changes in the rate of oxidation of CH_4 to H_2O in the upper stratosphere

* Polar Ozone and Aerosol Monitor.

† Microwave Limb Sounder.

may have occurred between 1960 and 1980. Just such a change in oxidation rate was observed by satellite between 1992 and 1999.

Future work should concentrate on more detailed calculations with a general-circulation model with full radiation scheme and including CH_4 chemistry, PSC formation, midlatitude ozone loss and ozone hole reactions. Calculations with such a model would allow the qualitative explanation given here to be tested quantitatively.

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